The Effect of Subtropical Cooling on the Amplitude of ENSO: A Numerical Study

De-Zheng Sun, Tao Zhang, and Sang-Ik Shin

NOAA-CIRES/Climate Diagnostics Center, Boulder, CO

Abstract

The effect of an enhanced subtropical surface cooling on ENSO through the "ocean tunnel" is

investigated using a coupled model. The "ocean tunnel" here refers to the water pathway that

connects the equatorial upwelling water to the subtropical/extratropical surface water. The

subtropical cooling is introduced through a reduction of the radiative convective equilibrium SST

(SST_p) in that region. The SST_p for the equatorial region is kept fixed.

It is found that an enhanced cooling in the subtropics results in a regime with stronger ENSO.

This is because an enhanced subtropical cooling reduces the temperature of the water feeding the

equatorial undercurrent through the "ocean tunnel". The resulting larger difference bewteen the

warm-pool SST and the temperature of the equatorial thermocline water —the source water for

the equatorial upwelling—tends to increase the equatorial zonal SST contrast between the

western and the eastern Pacific. In response to this destabilizing forcing to the coupled equatorial

ocean-atmosphere, stronger ENSO develops. ENSO is found to regulate the time-mean

difference between the warm-pool SST and the temperature of the equatorial undercurrent. The

findings provide further support for the "heat-pump" hypothesis for ENSO which states that

ENSO is an instability driven by the meridonal differential heating over the Pacific Ocean and

that ENSO regulates the long-term stability of the coupled equatorial Pacific climate. The results

also substantiate the notion that surface variability from higher latitudes may influence equatorial

SST variability through the 'ocean tunnel".

1.Introduction

The effect of ENSO on the subtropical/extratropical climate has been extensively studied and well understood (see review by Trenberth et al. 1998 and by Alexander et al. 2002). In contrast, our understanding of the influence of subtropical/extratropical climate on ENSO is very limited. This study is motivated to improve this situation, specifically by addressing the question how ENSO amplitude may respond to an enhanced subtropical cooling over the Pacific Ocean.

A significant role of the subtropical cooling in modulating the amplitude of ENSO has been hinted by the finding of an "ocean tunnel" –the water constituting the equatorial undercurrent and therefore the upwelling water in the equatorial Pacific comes from the subtropical/extratropical region (Pedlosky 1987, Liu et al. 1994, McCreary and Lu 1994). Gu and Philander (1997) even hypothesized that such a "tunnel" may be an underlying cause for the Pacific decadal variability. Zhang et al. (1998) provided further observational support for this hypothesis. Although it is found later that the processes in the real ocean appear to be more complex than what Gu and Philander (1997) originally envisioned (Lu et al. 1998, Schneider et al. 1999), the finding remains robust that the water feeding the equatorial undercurrent comes from the subtropics. Using two ocean GCMs, Shin and Liu (2001) demonstrated that a change in the subtropical/extratropical SST could induce a significant change in the temperature of the equatorial thermocline. Unfortunately, in their perturbation experiments, the equatorial winds are not coupled to the SST, effectively eliminating a necessary mechanism supporting ENSO. Nonetheless, their results add weight to the possibility of a significant subtropical/extratropical impact on ENSO through the aforementioned "ocean-tunnel". It has been known since the

pioneering study of Zebiak and Cane (1987) that El Niño results from instability of the coupled tropical ocean-atmosphere (see also the review by Neelin et al. 1998). A change in the equatorial thermocline temperature is an effective way to perturb the stability of the equatorial ocean. For example, a colder equatorial undercurrent would result in colder upwelling water in the eastern equatorial Pacific. The subsequent cooling on the SST in the eastern equatorial Pacific then strengthens the equatorial zonal SST contrast between the western and eastern Pacific, which may render the coupled equatorial Pacific system less stable, triggering the onset of a regime with stronger ENSO.

The need to examine the response of ENSO to an enhanced subtropical/extratropical cooling is further highlighted by the "heat-pump" hypothesis for ENSO (Sun 2003). Based on an analysis of the heat balance in the equatorial Pacific, Sun has suggested that the amplitude of ENSO may be linked to the poleward heat transport out of the equatorial ocean and therefore by extension to the meridional differential heating over the Pacific ocean. Sun (2003) has further carried out numerical experiments demonstrating that an increase in the equatorial surface heating (or equivalently an increase in the tropical maximum SST) results in a regime with more energetic ENSO. Specifically, Sun has found that an increase in the radiative heating over the equatorial region increases the difference between the tropical maximum SST—the warm-pool SST—and the temperature of the equatorial thermocline water. This further tends to increase the equatorial zonal SST contrast. Stronger ENSO develops to counteract this tendency. The difference between the warm-pool SST and the temperature of the equatorial thermocline water may also be perturbed by cooling the equatorial undercurrent which we expect an enhanced subtropical/extratropical cooling will induce via the aforementioned "ocean tunnel".

2. Methods

This study therefore aims to extend the study of Sun (2003) and to further explore the concept

that the ENSO is driven by the meridional surface differential heating over the Pacific. The same

numerical model used in that study will be used to investigate the effect of a cooling over the

subtropical Pacific on the amplitude of ENSO. The atmospheric component of the coupled model

of Sun (2003) is an empirical atmospheric model. The surface heat flux into the ocean is assumed

to be proportional to the difference between the radiative-convective equilibrium SST (SST_p) and

the actual SST,

$$F_s(\lambda, \phi) = C_p \rho c H_m (SST_p(\phi) - SST(\lambda, \varphi)) \qquad (1)$$

Where F_s is the net surface heat flux into the ocean, λ is the longitude, ϕ is the latitude, C_p is

the specific heat, ρ is the density, c is the restoring coefficient, and H_m is the depth of the mixed

layer (50m). The SST_p in the equation is prescribed empirically such that the model ocean is

heated in the equatorial region and cooled in the higher latitudes (see Eq. (5) in Sun (2003) for

the exact form of the prescribed SST_p). The restoring coefficient c is estimated from

observations (5.8 x $10^{-8}s^{-1}$) (Sun 2003). Over the equatorial region, the zonal wind is coupled to

the zonal SST gradients in the following form,

$$\tau^x(\lambda, \phi) = \tau^x_{ref}(\lambda, \phi) - \mu(\phi)(\Delta T - \Delta T_{ref}) \qquad (2)$$

Where τ^x is the zonal wind stress, τ^x_{ref} is the zonal wind stress that is used to spin up the ocean

model to obtain a reference state. The annual mean wind stress from the NCEP reanalysis for the

period of 1980-99 was used as τ^x_{ref}. ΔT_{ref} is a measure of the equatorial zonal SST contrast of the

reference state. It is defined as the area averaged SST difference between (5^oS-5^oN, 130^oE-

180°E) and (5°S-5°N, 230°E-280°E). ΔT is the same measure of the actual equatorial zonal SST

contrast predicted by the model. μ measures the coupling strength and has a prescribed

meridional profile—it decreases away from the equatorial region following a cosine profile.

Outside the region of 15°S-15°N, the influence of the equatorial zonal SST contrast on the winds

is assumed to be negligible. The value of μ at the equator, taken as 0.0060 $Nm^{-2}K^{-1}$, is assessed

from the observed interannual variations of zonal winds (NCEP winds) and the zonal SST

contrast. For example, by linearly regressing the monthly mean interannual anomalies of the

NCEP zonal wind stress (Kalnay et al. 1996) in the central equatorial Pacific (5°S-5°N, 160°E-

200°E) over the period 1980-2002 to the corresponding variations in the equatorial zonal SST

contrast (ΔT) (Rayner et al. 1996) , one obtains a regression coefficient of $0.0059 \pm 0.00053 Nm^{-2}K^{-1}$.

The ocean component is the NCAR Pacific basin model (Gent and Cane 1989, Gent 1991). The

ocean model is a primitive equation model for the equatorial upper ocean. It employs the reduced

gravity assumption so that the deep ocean is at rest. The model consists of a mixed layer and an

active layer below which is further divided into 6 numerical layers by means of sigma coordinate.

The temperature of each layer is calculated from its corresponding thermodynamic equation.

Therefore, different from the model of Zebiak and Cane (1987), the present ocean model

explicitly calculates the heat budget of the entire upper ocean. The model has 1° resolution in the

longitudinal direction and a variable one in the meridional direction. Within the equatorial

waveguide, the resolution is as fine as 0.25 ^{o}, thus ensures accurate simulation of the equatorial

waves and the associated dynamic adjustment.

Enhanced cooling in the subtropical ocean is introduced by reducing SST_p in that region. The dynamical coupling between the atmosphere and ocean in the coupled model is at present limited to the equatorial region. Coupling between the winds and the SST in the equatorial region is a necessary mechanism supporting ENSO (Zebiak and Cane 1987, Jin 1996, Neelin et al. 1998). The goal of the present effort is to investigate whether a cooling in the subtropical ocean can affect the amplitude of ENSO through the "ocean tunnel" that connects the surface ocean in the subtropical region with the subsurface ocean in the equatorial region (Pedlosky 1987, Liu et al. 1994, McCreary and Lu 1994).

3. Results

Fig.1a shows the time series of Niño3 SST from a control run and a perturbed run. The perturbed run has an enhanced cooling in the subtropical ocean. The exact difference between the SST_p in the control run and the perturbed run is shown in Fig.1b. To be in line with the heating experiment reported in Sun (2003), the enhanced cooling starts from the off-equatorial region and is assumed to increase with the latitude. The maximum cooling is reached at the poleward boundary of the ocean model: 30°S and 30°N. The working hypothesis is that enhanced subtropical/extratropical cooling affects the amplitude of ENSO through its effect on the temperature of the equatorial undercurrent (Sun 2000, Sun 2003). Therefore we are here simply seeking a cooling that will induce significant reduction in the temperature of the equatorial thermocline water. As we will see, such a form of cooling does cool the equatorial undercurrent significantly.

The two runs shown in Fig.1a start from the same initial condition. After some years of delay (about 8 years in this particular case), stronger ENSO develops. Note that La Niña becomes colder and El Niño become warmer. Comparing with the amount of cooling introduced outside of the equatorial region, the change in the amplitude of ENSO appears to be large. We will return to this sensitivity issue later. Fig. 2 further shows the differences in the equatorial upper ocean temperature between the control run and the perturbed run during La Niña, El Niño, and in the time mean. The subsurface temperature differences during La Niña and El Niño have a similar pattern to the corresponding temperature anomalies in the control run (Fig.3ab). The subsurface signatures of ENSO in both the western and the eastern Pacific are amplified by the imposed subtropical surface cooling. Therefore, we here present a good example showing how an external thermal forcing projects its impact on the internal mode of the system. The impact of this cooling on the time-mean state is much subdued, compared to the impact on the temperature structure during the two phases of ENSO.

The response of ENSO amplitude to the imposed subtropical cooling may be understood by first checking the effect of the subtropical cooling on the stability of the equatorial Pacific ocean-atmosphere. To this end, we have run the same experiments shown in Fig. 1, but without the coupling between the equatorial surface winds and the equatorial zonal SST gradients. Therefore, the feedback from the coupled instability—ENSO—is artificially suppressed. Fig. 4a compares the value of the temperature of the equatorial undercurrent (T_c) from the control experiment with that from the perturbed experiment as a function of time. T_c is defined here as the zonal mean core temperature of the equatorial undercurrent. It is obtained by first finding the equatorial

ocean temperature at the depth where the zonal velocity has its maximum value and then averaging this temperature across the region of 180°E-270°E. Tc is meant to be a simple measure of the characteristic temperature of the equatorial thermocline water. Fig.4a shows that the value of Tc from the perturbed experiment with enhanced subtropical cooling starts to diverge from the corresponding value of Tc from the control run after a few years. Fig. 4b and Fig. 4c further show the meridional cross sections in the central (160°E-210°E) and the eastern Pacific (210°E-270°E) of the upper ocean temperature differences between the control run and the perturbation run. The surface cooling in the subtropical ocean clearly sinks into the equatorial thermocline. As a consequence, the equatorial thermocline water is about 1 ^{o}C colder across the entire equatorial Pacific (Fig. 5). The temperature anomaly in the equatorial thermocline is even larger than the surface cooling imposed in the subtropics. This amplification may have something to do with propagation of Rossby waves originating from the forcing region into the baroclinically unstable regions in the lower limb of the subtropical gyre. Note that in this model, there is no salinity. The temperature perturbations are also density perturbations. Liu and Shin (1999) show that the subduction of density anomalies in their model may be explained as propagation of baroclinic Rossby waves. Galanti and Tziperman (2003) suggest that the equatorward propagating Rossby waves could be amplified in passing through the baroclinically unstable regions in the lower limb of the subtropical gyre.

Without the equatorial ocean-atmosphere coupling, the 1 ^{o}C or so cooling to the thermocline water causes little change in the SST of the western Pacific warm-pool (Fig. 5). The mean SST over the equatorial western Pacific warm-pool region (5°S-5°N, 120°E-160°E)—we here use Tw to represent this temperature—is reduced only by 0.04 ^{o}C .   Two factors apparently prevent

significant cooling to the warm-pool SST. One is the considerable depth of the thermocline in the

western Pacific. The other is the fact the radiative convective equilibrium SST (SST_p) is kept

fixed for the equatorial region (Fig.1b). Therefore, the imposed subtropical cooling increases

considerably the difference between the warm-pool SST and the temperature of the equatorial

thermocline water (the value of Tw-Tc is increased by about 1 ^{o}C).

The difference between Tw and Tc is a key parameter determining the stability of the coupled

equatorial ocean-atmosphere. The larger the difference between Tw and Tc, the less stable the

coupled system (Sun and Liu 1996, Sun 1997). This is because the value of Tw-Tc affects the

zonal SST contrast and thereby the zonal wind stress. A positive perturbation to the equatorial

zonal SST contrast is evident in Fig. 5. The shallower thermocline in the central and eastern

Pacific apparently allows the colder equatorial upwelling water to cool the SST in that region

more effectively than it does to the western Pacific SST. Measured by ΔT in Eq. (2), the

equatorial zonal SST contrast in the absence of ocean-atmosphere coupling is increased by 0.27

^{o}C. This increase is small, but significant. Once the equatorial zonal wind stress is coupled to the

zonal SST gradients via Eq. (2), this initial cooling to the eastern Pacific SST is amplified:

stronger winds result in stronger upwelling which in turn strengthens the zonal SST contrast. As

shown in Fig.2c, the zonal SST contrast in the cold period of the coupled case is increased by as

much as 1.9 ^{o}C. Therefore, the subtropical cooling impinges on the stability of the coupled

equatorial Pacific ocean-atmosphere system via its effect on the difference between Tw and Tc

and then on the equatorial zonal SST contrast. The effect on the latter involves the coupling

between the atmosphere and ocean. Apparently, the onset of a regime with stronger ENSO is a

coupled response to this destabilizing tendency. Extending this reasoning, one expects that the

ENSO may act as a stabilizer of the coupled equatorial Pacific ocean-atmosphere system: the presence of ENSO may reduce the sensitivity of Tw-Tc in the mean state to an external forcing.

Indeed, once the coupling is turned on and ENSO is allowed to develop, the enhanced subtropical cooling has a much reduced impact on the difference between Tw and Tc in the time-mean state. Comparing Fig. 5 with Fig. 2c, it reveals that the change in the time-mean temperature of the equatorial thermocline water in the coupled case is less than half of that in the uncoupled case. (The cooling to Tc is reduced from about 0.85 ^{o}C to 0.32 ^{o}C). Also, ENSO causes significant cooling to the SST of the western Pacific warm-pool. (Tw is cooled by 0.79 ^{o}C in the coupled case compared to a negligible 0.04 ^{o}C in the uncoupled case). Consequently, the difference between the warm-pool SST and the temperature of the equatorial thermocline water induced by the imposed subtropical cooling in the coupled case (Fig.2c) is much reduced compared to that in the corresponding uncoupled case in (Fig. 5). The lack of change in the time-mean difference between the warm-pool SST and the temperature of the equatorial themocline water also prevents significant change from occuring in the equatorial zonal SST contrast. Measured by the value of ΔT in Eq. (2), the change in the time mean zonal SST contrast is a negligible 0.025 ^{o}C. The apparent lack of sensitivity in the time-mean zonal SST contrast even in the presence of the positive Bjerknes feedback is also linked to the fact that the enhanced subtropical/extratropical cooling apparently mainly cools the cold phase of ENSO in the equatorial region east of the dateline (Fig. 6b). But then El Niño events are stronger, which largely reverses the cooling effect during the cold phase (Fig. 6c). This cancellation effect is evident in Fig.2.

Detailed examination of the coupled perturbed run reveals that the onset of the regime with

stronger ENSO is triggered when the equatorial thermocline water is cooled significantly and when this cooling coincides with an on-going La Niña event. Fig. 7 shows the time series of the temperature of the equatorial undercurrent (Tc) for the coupled control run and the corresponding perturbed run. A significant cooling to Tc occurred at about the 44th month, but this cooling is short-lived. This cooling may indicate the passage of the first packet of waves. A persistent difference between the Tc in the perturbed run and the Tc in the control run starts to occur at about the 55th months. This is followed by a significant difference between the Niño3 SST in the perturbed run and the control run about 5 months later (see Fig.1a). This initial cooling happens to be in a growth phase of El Niño. This El Niño is therefore damped. The subsequent La Niña is, however, strengthened by this cooling. After this stronger La Niña, El Niño becomes warmer, apparently because of the coupling. A regime with stronger ENSO then develops.

To confirm the order of this sequence of events, we have run another pair of experiments—a coupled control run and a corresponding perturbed run. The Niño3 SST and Tc from these two runs are shown in Fig. 8. This pair of runs only differs from the previous pair in the initial conditions from which the integration starts. Again, the imposed surface cooling triggers the onset of a regime with stronger ENSO at the time when its cooling effect on the equatorial thermocline coincides with an ongoing La Niña.   Note also that Tc generally leads the Niño3 SST though the lead varies (2-8 months). Tc is anomalously warm prior to the onset of the warm events and anomalously cold prior to the onset of the cold events.

El Niño in this model results from a coupled instability. The reason that a stronger El Niño

develops after a stronger La Niña is because a stronger La Niña pumps more heat down to the

subsurface ocean of the equatorial western Pacific and renders the coupled system more unstable

(Fig. 2a). ENSO in this model behaves like a pendulum: if you push the pendulum to the

direction where it is already swinging into, it will be able to climb higher when it swings back.

Stronger El Niño results in stronger poleward heat transport. Figure 9 shows the time series of

the poleward heat transport away from the equatorial Pacific from the control run and the

perturbed run. Comparing with Fig. 1, it shows that the poleward heat transport lags the surface

warming: as the surface warming reaches its peak, the equatorial Pacific continues to lose more

heat to the higher latitudes, pushing the equatorial Pacific to a La Niña state. This lag between

the surface warming and the poleward heat transport therefore acts as a phase transition

mechanism.

Fig.1 and Fig.8 also show that in response to the subtropical cooling, the period of the oscillation

becomes longer. This reduction in the frequency of ENSO is probably linked to the increase in

the amplitude of ENSO. As originally found in the model of Zebiak and Cane (1987) and further

delineated in the weakly non-linear analysis of Jin (1997), a stronger oscillation tends to be

accompanied by a reduced frequency—a longer period. Detailed analysis of factors that influence

the frequency of ENSO in the present model will be presented in a later paper.

The sensitivity of ENSO in the model to the subtropical cooling appears to be high. The cooling

to the equatorial thermocline water is about 1.0 $^{\circ}$C (Fig.4 and Fig.5). The change in the amplitude

of ENSO is from about 2 $^{\circ}$C to about 6 $^{\circ}$C: the Niño3 SST at the peak of the La Niña phase is

about 3.5^{o}C cooler while the Niño3 SST at the peak of the El Niño phase is about 0.5 ^{o}C warmer (Fig. 1a). The amplification of the cooling during the La Niña phase is due to the Bjerknes feedback—the positive feedback loop between the equatorial zonal SST gradients, the equatorial zonal wind stress, and the equatorial upwelling. An initial cooling to the eastern equatorial Pacific SST due to the upwelling of the colder equatorial thermocline water causes a strengthening of the zonal SST contrast and thereby a strengthening of the zonal winds. The slightly stronger El Niño is a simple response to the stronger La Niña as the two phases of ENSO involve a zonal redistribtuion of the heat in the equatorial upper ocean. Therefore the sensitivity of ENSO to the subtropical cooling depends on the strength of the Bjerknes feedback . The strength of the Bjerknes feedback is measured by the coupling strength parameter μ. As mentioned before, our estimate of μ is based on the NCEP reanalysis, and therefore is only realistic to the degree that the NCEP wind stress is realistic. In addition, the coupling between the winds and the SST in the present model is also idealized. It is possible that in a more sophisticated model, the sensitivity of ENSO to subtropical cooling is weaker. On the other hand, we have to note that the imposed cooling is not small—it is certainly larger than observed changes in the subtropical Pacific temperature on decadal time-scales. For example, the maximum difference in the extratropical Pacific SST between the epoch 1977-93 and the preceding epoch 1950-1976 is less than 0.5 ^{o}C (Zhang et al. 1997).

We have further verified that the delay in the response of ENSO to a subtropical/extratropical cooling become progressively longer as we move the cooling progressively to the higher latitudes. An example is shown in Fig. 10. Now the onset of a regime with stronger ENSO does not occur until almost 12 years later.

We have also found that the cooling outside the region of 20^oS--20^oN contributes to most of the response of ENSO in Fig.1. Fig.11a and Fig.11b shows respectively the response of Niño3 SST to the surface cooling shown in Fig.11c and Fig.11d. The surface cooling used for Fig.1 is now divided into two parts. Cooling shown in Fig. 11d is confined to the region poleward 20°. Fig. 11c shows the remaining part of the cooling. The initial response in the Niño3 SST variability shown in Fig.1a is apparently due to the cooling in the region of 5^oN(S)--20^oN(S), but the influence of cooling from this region on the Niño3 SST is weak and is dampened after 10 years of its arrival. The response in amplitude due to the cooling in Fig. 11d is very close to the response to the total cooling shown in Fig. 1. This is because the cooling to the equatorial thermocline water shown in Fig. 5 is mainly due to the surface cooling imposed in the region poleward of 20^oS--20^oN. Fig. 12a and Fig. 12b show respectively the effect of the surface cooling shown in Fig. 11c and Fig.11d on the equatorial thermocline temperature.

We have also done experiments with the surface cooling replaced by surface warming. With sufficient strong subtropical warming, ENSO can be killed entirely (Fig. 13). This result is expected from the cooling experiments. While subtropical cooling tends to destabilize the equatorial ocean-atmosphere system through reducing the temperature of the equatorial thermocline water, the subtropical heating tends to stabilize the equatorial ocean-atmosphere system by warming up the equatorial thermocline water. The warming effect of the surface heating shown in Fig. 13b on the equatorial thermocline water is shown in Fig.14. To isolate the stabilizing effect of the subtropical heating on the equatorial ocean-atmosphere system, the equatorial ocean-atmosphere coupling is again turned off in obtaining Fig.14. The figure shows a

substantial reduction in the value of Tw-Tc (~1.6 ^{o}C) and a substantial reduction in the value of ΔT (~0.56 ^{o}C). Consequently, the equatorial ocean-atmosphere system is completely stabilized.

The domain of the present model prevents us from moving the surface cooling beyond regions poleward 30°N or 30°S. As we remarked before, the ideal setting to test the heat-pump hypothesis is to impose a cooling that centers about the latitude where the 20°C isotherms outcrops. For the observed annual mean situation, this latitude is about 35°N in the north Pacific and 30°S in the south Pacific. This is why we maximize the cooling near the model boundary in the present model setting. We are also helped by the fact that surface cooling imposed in the subtropical region sinks locally to considerable depth (Fig.4bc). Provided that surface cooling in these regions induces upper ocean temperature differences within the tropical domain similar to those shown in Fig.4bc and Fig.5, we do not expect any qualitative difference in the effect of an mid latitude cooling on ENSO from the experiments we have reported here. Therefore, the cooling from the higher latitudes via subduction may be partially simulated here.

A related question is whether ENSO will still respond the same way when surface cooling is imposed in the very high latitudes where the intermediate water and deep water originate. Considering that such a cooling will ultimately affects the equatorial deep ocean temperature via the global thermohaline circulation (Broecker 1987), Sun (2000) suggests a positive impact of such a cooling on ENSO, citing results from a highly idealized model.   By simply reducing the equatorial deep ocean temperature in the present model, we find that a cooling in the deep ocean temperature in this model indeed results in a regime of stronger ENSO (Fig.15). While this result supports the speculation by Sun (2000), it should be viewed with caution as the ocean model

used in this study was mainly designed as an upper ocean model, and the communication

between the equatorial deep ocean and the equatorial upper ocean in this model is not realistic.

For example, a reduction in the deep ocean temperature in the present model is immediately felt

by the upper ocean overlying it via entrainment (Gent and Cane 1989). This treatment may not

simulate the actual upwelling from the deep ocean.

4. Conclusions

Numerical experiments have been conducted to investigate the effect of an enhanced subtropical

surface cooling on the level of ENSO activity. The results show that such a perturbation to the

surface heating can indeed result in a regime with stronger ENSO activity. Moreover, ENSO

appears to regulate the time-mean difference between the warm-pool SST (Tw) and the

temperature of the equatorial undercurrent (Tc).

The enhanced subtropical surface cooling influences the level of ENSO through cooling the

equatorial undercurrent and the associated equatorial upwelling water. With the radiative

convective equilibrium SST over the equatorial region being kept fixed, the difference between

the warm-pool SST (Tw) and the temperature of the equatorial undercurrent (Tc) tends to

become larger. This tendency constitutes a destabilizing effect on the stability of the coupled

equatorial ocean-atmosphere system. A regime with stronger ENSO develops apparently as a

coupled response to this destabilizing forcing. While the enhanced subtropical cooling is found

to induce a substantial increase in the difference between Tw and Tc in the absence of ENSO, it

results in little increase in the time-mean difference between Tw and Tc in the presence of

ENSO. This contrast in the response to the enhanced subtropical cooling provides evidence that

ENSO regulates the long-term stability of the coupled equatorial ocean-atmosphere system.

The findings here are consistent with the argument presented in Sun (2003): El Niño corresponds to a poleward heat transport mechanism; Either an increase in the equatorial surface heating or an increase in the subtropical/extratropical cooling should result in a regime with more energetic ENSO events. Moreover, ENSO may be a mechanism that regulates the long-term stability of the coupled equatorial ocean-atmosphere system. In addition, the present findings appear to substantiate the argument of Gu and Philander (1997) and Zhang et al. (1998) that the "ocean tunnel" may be an important mechanism for decadal variability.

Surface thermal forcing in the subtropical/extratropical region may affect equatorial SST variability through many pathways. The present study focuses on the role of the "ocean tunnel". Correspondingly, the experiments reported here are highly idealized. The winds and SST are only coupled in the equatorial region. Presumably, as the subtropical/extratropical ocean cools, the Hadley circulation may become stronger and the subtropical/extratropical surface winds may change correspondingly. These changes may in turn influence the equatorial SST response. This model also lacks a seasonal cycle and the associated monsoonal circulation, both of which have been hypothesized as important players in projecting the influence of a global forcing over ENSO (Clement et al. 2002, Liu et al. 2000). Our next step is to couple the present ocean model with a global atmosphere so that the feedbacks from the global atmospheric circulations could be better represented. The present results should be useful for understanding results from more complicated models, such as those from a recent study by Otto-Bliesner et al (2003) using NCAR CSM.

Additional experiments with a global ocean model are also needed. A global ocean model allows us to examine more carefully the response of ENSO to enhanced cooling in the region poleward of 30^oN and 30^oS including cooling in other basins. A global ocean model may also reveal a role of the global thermohaline circulation in linking ENSO to the high latitude surface conditions. All these details may potentially modify the mechanism suggested here by which subtropical/extratropical cooling affects ENSO. Nonetheless, the present experiments are potentially valuable as a step stone for a more complete understanding and assessment of the role of the "ocean tunnel" in linking the high latitude surface forcing to the behavior of ENSO on decadal and longer time-scales.

The ultimate test of the mechanism by which a subtropical/extratropical cooling on ENSO and the heat-pump hypothesis for ENSO in general, has to come from observations, paleo-climate observations in particular. In this connection, it may be worth noting the recent data from Cobb et al. (2003). Her coral records appear to suggest that ENSO during the little ice age (LIA) was stronger than in the medieval warm period (MWP). She has also noted in the same record that the time-mean tropical climate has little change from periods of weak ENSO activity to periods of strong ENSO activity. The present findings concerning the regulatory effect of ENSO on the tropical mean climate suggest that the lack of change in the local time-mean state (i.e., the mean tropical Pacific climate) may be precisely due to the large variations in the level of ENSO activity. Chaos and weather noise within the tropics may play a significant role in the variability of ENSO on decadal and longer time-scales. But the present results suggest that in order to identify all the causes for the variability of ENSO on decadal and longer time-scales, one may

have to keep in mind the role of the "ocean tunnel" and look beyond the tropical Pacific and look

poleward.

Acknowledgments

Figure Legends

Figure 1: (a) Time series of Niño3 SST from the control run (solid line) and the perturbed run
(dashed line). The two runs start from the same initial condition. The perturbed run is
subjected to a reduction in SST_p in the form shown in (b).

Figure 2: The differences in the equatorial upper ocean temperature (5^oS-5^oN) between the
coupled control run and the coupled perturbed run (perturbed run minus control run) in
the cold phase (a), the warm phase (b), and the time-mean (c). The definition of the warm
and cold phases is the same as in Sun (2003). The last 16 years of the run shown in Fig.1a
are used in this calculation.

Figure 3: The equatorial upper ocean temperature anomalies during the cold phase (a), and the
warm phase (b) of the control run whose Niño3 SST is shown in Fig.1a.

Figure 4: (a) Time series of the core temperature of the equatorial undercurrent (Tc) from two
uncoupled runs. The solid line corresponds to the control run while the dashed line corre-
sponds to the perturbed run. The form of cooling that the perturbed run is subjected to is
the same as the one for the coupled perturbed run (Fig.1b). The value of Tc is obtained by
first finding for all longitudes the equatorial ocean temperature at the depth where the
zonal velocity has its maximum, and then zonally averaging this temperature from 180^oE
to 270^oE . (b) &(c): Meridional across sections of the upper ocean temperature difference
in the central (160^oE-210^oE) and eastern Pacific (210^oE- 270^oE) during the last three years

of the run. The thin dashed contours indicate the mean isentropes of the control run.

Figure 5: The temperature differences in the equatorial upper ocean (5^oS-5^oN) between the uncoupled control run and the corresponding uncoupled perturbed run described in Fig.4 bc. As for those figures, the last three years of a 23-year long run are used in the calculation.

Figure 6: Zonal distribution of the equatorial SST (5^oS-5^oN) for the control run (solid line) and the perturbed run (dashed line) in the time mean (a), during the cold phase (b), and the warm phase (c).

Figure 7: Time series of the core temperature of the equatorial undercurrent (Tc) from the coupled control run (solid line) and the corresponding perturbed run (dashed line). The value of Tc is obtained the same way as in Fig. 4a.

Figure 8: Time series of Niño3 SST (a) and Tc (b) from another pair of coupled runs-- a control run (solid line) and a perturbed run (dashed line). This pair of runs is the same as the pair of runs shown in Fig.1 and Fig.7 except in the initial conditions from which the runs start. The value of Tc is calculated the same way as in Fig.4a.

Figure 9: Time series of the poleward heat transport out of the equatorial Pacific (5^oS-5^oN) from the control run (solid line) and the perturbed run (dashed line).

Figure 10: (a) Time series of Niño3 SST from a control run (solid line) and a perturbed run (dashed line). The control run is the same as the control run shown in Fig.1a. The perturbed run shown here is subjected to a reduction in SST_p in the form shown in (b). The cooling starts from a higher latitude than that in Fig.1b. Note that with the cooling starting from a higher latitude, there is a prolonged delay in the response of the Niño3 SST.

Figure 11: (a) Response of Niño3 SST to cooling in the form shown in (c). (b) Response of Niño3 SST to cooling in the form shown in (d). Note that the sum of (c) and (d) equals to the cooling shown in Fig.1b.

Figure 12: (a) The cooling to the equatorial thermocline water due to the cooling shown in Fig.11c. (b) The cooling to the equatorial thermocline water due to the cooling shown in Fig.11d. Shown are the upper ocean temperature differences in the equatorial region (5°S-5°N) between an uncoupled control run and the corresponding uncoupled perturbed run. Both runs are 23 years long. The last three years of the runs are used for this calculation. The dashed contours indicate the mean isentropes of the control run.

Figure 13: (a) Response of Niño3 SST to subtropical heating. The solid line is for the control run. The dashed line is for the perturbed run. (b) The subtropical heating imposed in the perturbed run.

Figure 14: The equatorial upper temperature differences (5°S-5°N) induced by the subtropical

heating shown in Fig.13b. Shown are the mean differences between an uncoupled control

run and an uncoupled perturbed run over the last three years of a 23-year long run. Note

the significant reduction in the difference between western Pacific warm-pool SST and

the temperature of the thermocline water (Tw-Tc) and the corresponding reduction in the

zonal SST contrast (ΔT) between the equatorial western Pacific and the equatorial

eastern Pacific.

Figure 15: Time series of Niño3 SST from a control run (solid line) and a perturbed run (dashed

line). The control run is the same as the control run in Fig. 1a. The perturbed run has a

colder deep ocean temperature--2 K colder than in the control run. Recall that in this

model, there is only one temperature for the deep ocean--the part of the ocean that does

not feel the wind forcing and is motionless. The deep ocean temperature, however, affects

the upper ocean temperature by entrainment (Gent and Cane 1989).

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